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Thermal and mechanical differences between ice and water surfaces at an ice edge create an adjustment region. This adjustment region will be most "frontal" in character when the resulting gradients are largest, which is most probable when advection perpendicular to the ice edge is small. Thus a geostrophic wind nearly parallel to the ice edge gives much stronger surface temperature gradients (Kantha and Mellor, 1989) and stronger jet maxima (Langland et al., 1989) than when the geostrophic wind is perpendicular to the ice edge.

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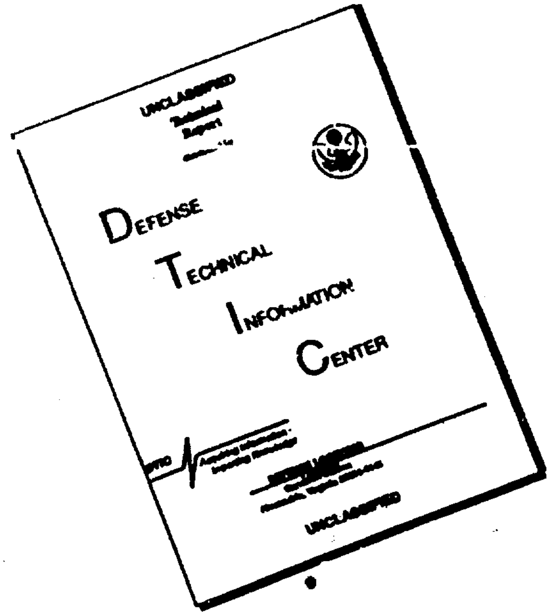

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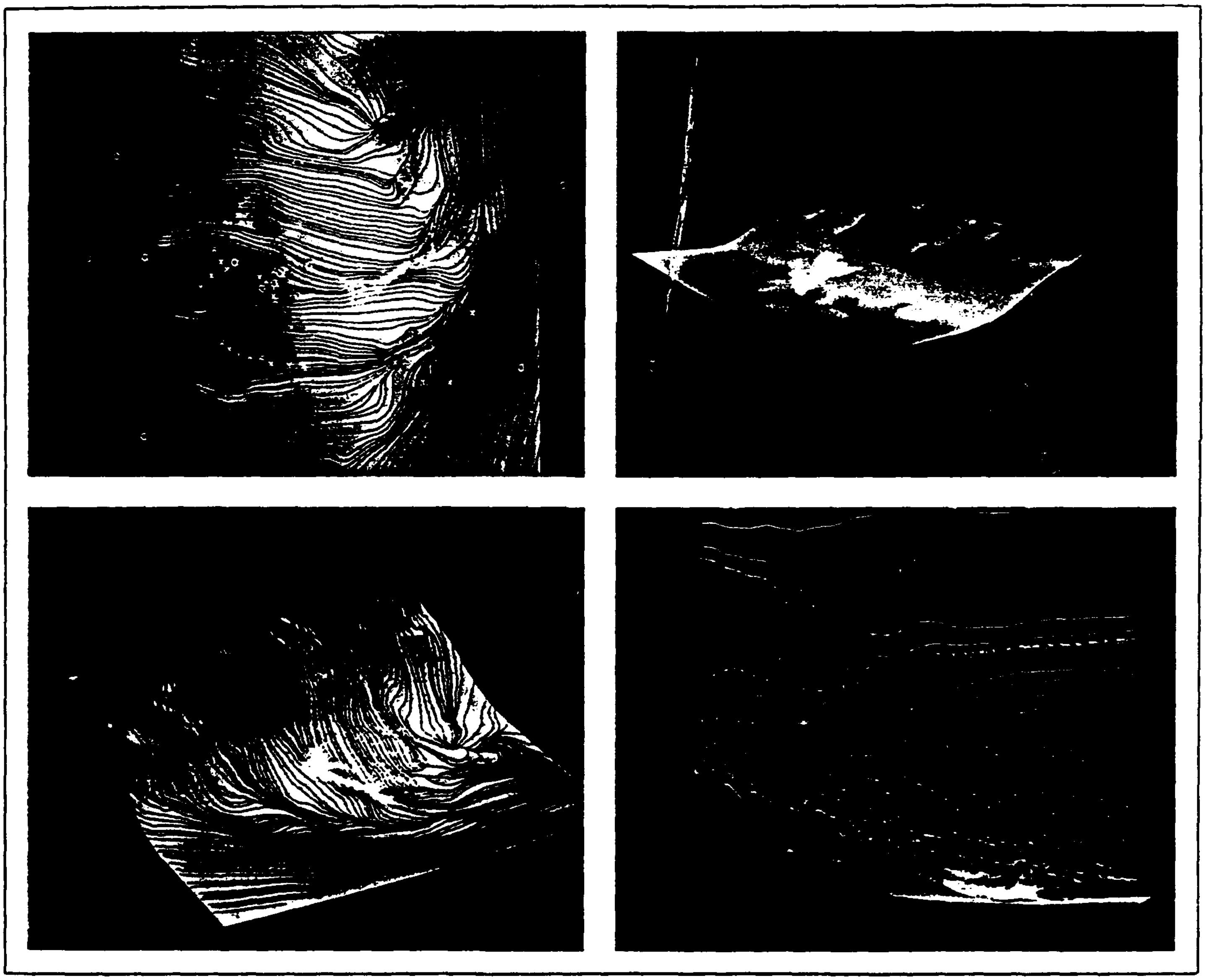

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BOUNDARY-LAYER STRUCTURE NEAR AN ICE EDGE AS A FUNCTION OF WIND DIRECTION

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1. INTRODUCTION

This paper examines the sensitivity of boundary-layer (BL) "frontal" features, created by differences across an ice edge and characterized by positive vertical velocity (w), to variations in geostrophic wind direction (α), particularly for flow approximately parallel to the ice edge. Notable is the asymmetry introduced by Coriolis/frictional forces.

Thermal and mechanical differences between ice and water surfaces at an ice edge create an adjustment region. This adjustment region will be most "frontal" in character when the resulting gradients are largest, which is most probable when advection perpendicular to the ice edge is small. Thus a geostrophic wind nearly parallel to the ice edge gives much stronger surface temperature gradients (Kantha and Mellor, 1989) and stronger jet maxima (Langland *et al.*, 1989) than when the geostrophic wind is perpendicular to the ice edge.

Neglecting friction, "frontal" features would be stronger for a geostrophic wind with an on-ice component than for one exactly parallel to an ice edge since on-ice advection from the geostrophic wind counters off-ice advection produced by the thermally induced surface flow. If the on-ice geostrophic component is too large, however, geostrophic advection would disrupt the thermally induced circulation and weaken gradients. Clearly there would be an optimal geostrophic wind angle which produces the strongest gradients.

But in addition, Coriolis forces couple the two components of horizontal momentum and create, with surface friction, an asymmetry in the optimal geostrophic angle between flows with ice on the "left" and on the "right". For an ice-parallel geostrophic wind with "ice on the left" (when looking downwind), frictional forces create an on-ice component to the wind, whereas for a parallel geostrophic wind with "ice on the right" the frictionally induced flow is off-ice. Thus a geostrophic wind with "ice on the right" requires a larger on-ice component to cancel the combined effects of thermal and frictional advection than would a geostrophic wind with "ice on the left". Similar arguments apply to an on-shore sea breeze (Estoque, 1962).

2. SIMULATION PARAMETERS

For an idealized linear ice-edge, the geostrophic wind angle α is rotated over a 360° range, using an orientation with ice on the "west" and water on the "east" where $\alpha=0°$ represents a wind from the "north". The x direction is perpendicular to the ice edge at x=0 and increases to the "east".

Modeling parameters are idealized to simplify interpretation of the results. The simulations assume the ice and water to have respective surface temperatures of -6°C and -2°C with surface roughnesses of 1cm and 0.005cm. The surrounding large-scale atmosphere stratification is $\partial\theta/\partial z=6$K/km and the geostrophic wind is 5m/s. (In this paper "geostrophic wind" will refer only to the large-scale forcing, not to pressure gradients created by the model.) A latitude of 79°N is assumed.

Beginning from the specified large-scale conditions, the simulation spin-ups to a quasi-equilibrium after a simulated 20 hours. The model employed is a two-dimensional hydrostatic mesoscale model employing a turbulent kinetic energy budget for BL parameterization, as described in Skupniewicz *et al.*, 1991. It employs finite-elements, allowing a stretched grid; the grid spacing varies with the geostrophic wind direction, having a typical domain of 260km horizontally and 2200m vertically with minimum horizontal resolution of 2km at the ice edge. A horizontal diffusivity coefficient of $500m^2/s$ is employed.

3. SENSITIVITY OF VERTICAL VELOCITY MAXIMUM

Fig. 1 depicts the maximum upward velocity —which occurs within the BL— and its location over a 360° range in geostrophic wind direction. The strong influence of geostrophic advection when the geostrophic wind is perpendicular to the ice-edge is evident, giving a relative minimum in maximum vertical velocity for α near 90° and 270°. This advection also produces relatively weak thermal gradients near the surface.

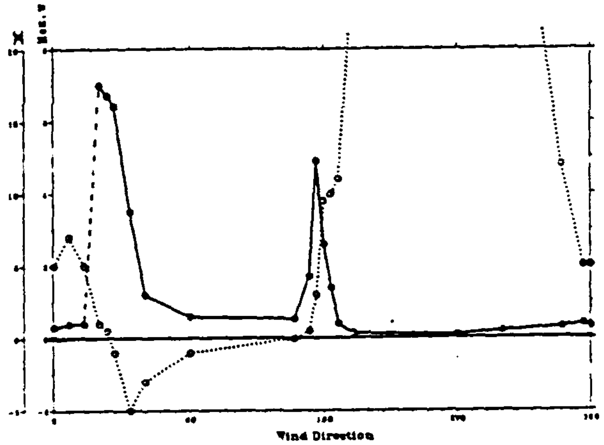

Fig. 1: Magnitude (solid, in cm/s) and position (dotted, in km) of domain maximum of vertical velocity as a function of geostrophic wind angle α.

The sensitivity to geostrophic wind direction is notable for flow nearly parallel to the ice with "ice on the left": the maximum vertical velocity is largest at $\alpha = 175°$ and is halved by a shift of 5°. (Note that the wind direction was incremented by 5°, so a somewhat higher maximum w might lie to one side of that for 175°.)

Fig. 2 depicts w contours for α of 180° and 175°. The additional on-ice geostrophic advection for 175° (Fig.2b) forces the maximum w closer the ice edge, increases its magnitude, steepens the BL slope over water, and increases the temperature gradient at the ice edge (not shown).

Now examining the w maximum with "ice on the right", for which frictional turning decreases the geostrophic wind advection, the asymmetry is dramatic. This case requires a 25° larger on-ice angle and its w maximum is 40% larger than that with "ice on the left". The sensitivity to wind direction is similar for both cases.

The "ice on the right" case requires elaboration, since the conditions which produce it are more complex than those for the "ice on the left". With the "ice on the right", as the on-ice angle of the geostrophic wind increases from 0° the upwind propagating transient —created as the model adjusts towards equilibrium— decreases in speed and becomes increasingly front-like. When the wind angle reaches 30°, the geostrophic "headwind", becomes large enough to check that feature near the ice edge: it then becomes the stationary w maximum. The dashed line in Fig. 1 represents this mode transition.

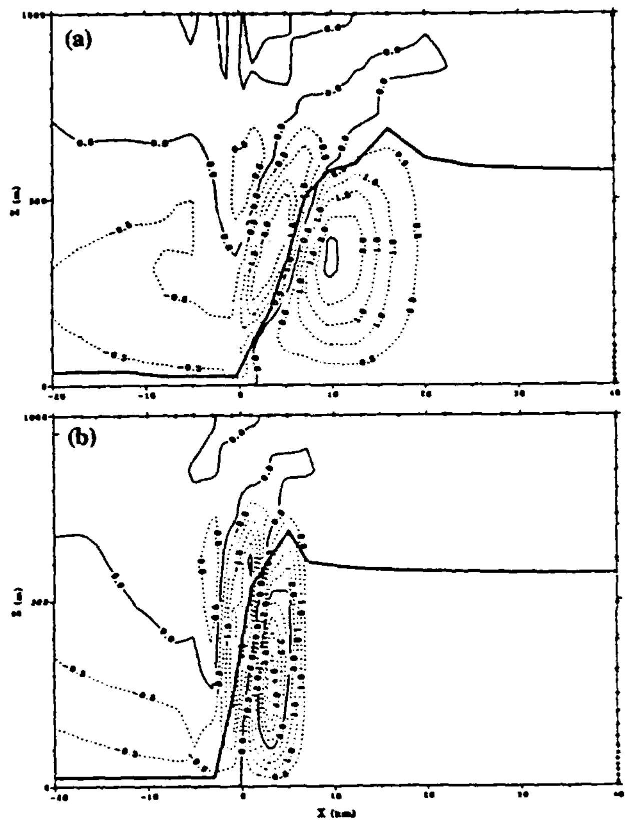

Fig. 2: Vertical velocity (in cm/s) for: (a) $\alpha = 180°$ and (b) $\alpha = 175°$. Thick solid line indicates BL depth, as diagnosed from turbulence kinetic energy equation.

Distinction between the desired stationary solution and the slowly propagating transient is validated both by extending the simulation time and by transient propagation analysis (which indicated that the transient feels an effective "headwind" which is 1/4 of the actual opposing geostrophic component). Strictly speaking this transient is a numerical result, a response to an initial model imbalance, but its existence suggests the physical effect that changes in wind direction for on-ice flow with "ice on the right" may create propagating disturbances in the BL. For example, decreasing the on-ice angle from 40° to 10° would allow the w maximum feature to propagate off-ice into the marine BL.

The asymmetry of the response is approximately 15°, i.e. the on-ice angles of maximum w are shifted 15° clockwise from being symmetric. This angle may be compared to the frictional shifts produced in a horizontally homogeneous BL over the specified ice and water surfaces: for a one-dimensional simulation, the BL-averaged velocity backs 18° from geostrophic over the ice surface and 4° over the ocean surface.

4. SENSITIVITY OF HORIZONTAL VELOCITY AND STRESS COMPONENTS

The sensitivity of the horizontal velocity field to small changes in wind direction for flow nearly parallel to the ice edge is presented in Fig. 3 and Fig. 4. Fig. 3, for the horizontal x component (u), illustrates how increasing the geostrophic "headwind" component, by rotating the geostrophic angle from 180° (Fig.3a) to 175° (Fig.3b), greatly reduces the region of positive u velocity near the surface. Note that the structure of the positive u velocity in Fig. 3a closely resembles the "head" feature observed in sea-breeze fronts.

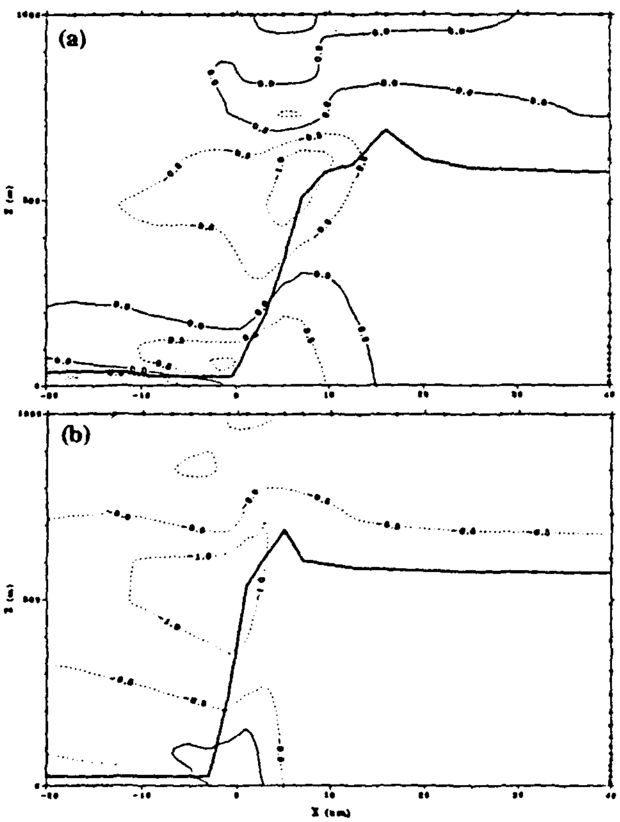

Fig. 3: Horizontal x velocity component (in m/s) for: (a) $\alpha=180°$ and (b) $\alpha=175°$.

Fig. 4, for the horizontal y component (v), illustrates the simultaneous shift in the positions of the jets created parallel to the ice-edge, one jet over the ice above the BL and another over the water within the BL. Increasing the on-ice component of geostrophic advection moves both jets "downwind". The distance between these jet positions depends principally upon, and can be predicted from, the large-scale stratification and BL depth over the ocean (Glendening, 1992).

Increasing the geostrophic headwind component beyond that of Fig. 3 and Fig. 4 destroys both the region of positive u velocity and the ice-edge-parallel jets.

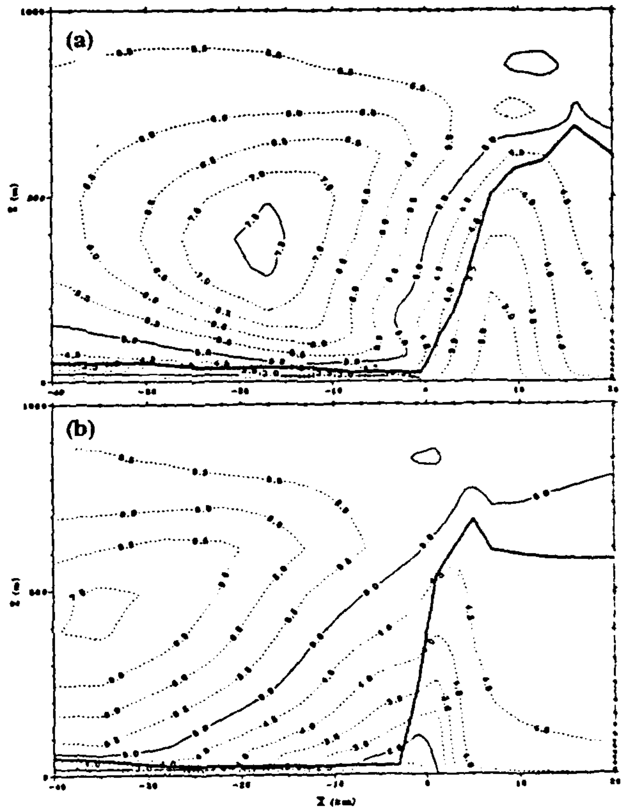

Fig. 4: Horizontal y velocity component (in m/s) for: (a) $\alpha=180°$ and (b) $\alpha=175°$.

The above velocity sensitivities are reflected in the surface stress variations of Fig. 5. Note that the dramatic change in the stress curl $\partial\tau_y/\partial x$ —including a sign change— over the water in Fig. 5a is absent in Fig. 5b, suggesting significantly different ocean responses for the 5° wind shift.

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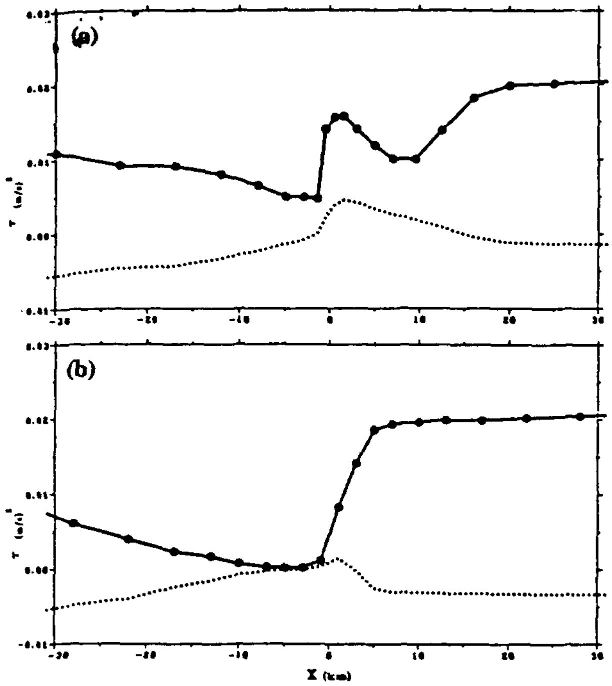

Fig. 5: Horizontal stress components (τ_x dotted, τ_y solid, in m^2/s^2) for: (a) $\alpha=180°$ and (b) $\alpha=175°$.

The above discussion assumes a Northern Hemisphere rotation. In the Southern Hemisphere, reversal of the terms "right" and "left" when describing the relationship between the ice edge and geostrophic wind direction is required.

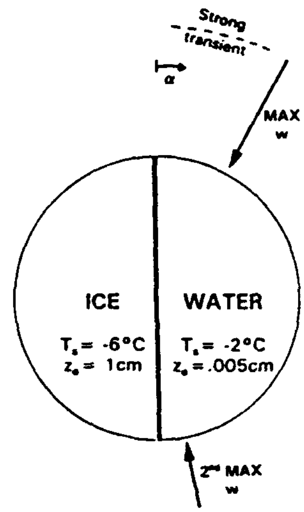

Fig. 6: Summary diagram for response as a function of geostrophic wind direction. Arrow length represents relative magnitude of maximum upward velocity.

5. CONCLUSIONS

The dependence of "frontal" features, created by thermal differences across an ice edge, upon geostrophic wind angle has been investigated. The principal findings are illustrated schematically in Fig. 6. The asymmetry resulting from Coriolis and frictional forces is apparent: for geostrophic flow with "ice on the right", a greater on-ice angle is required to obtain strong gradients perpendicular to the ice edge and the maximum upward velocity is larger than for the case with "ice on the left". The position of maximum w is closely related to its magnitude, since strong gradients at the ice edge require the maximum w to also occur there. The asymmetry is of the same magnitude as the frictional turning expected for a horizontally homogeneous BL. The asymmetry also appears through the creation of strong transients for flow with "ice on the right", suggesting that small changes of wind direction for such a condition may create propagating disturbances in the BL.

Other variables are equally affected by this advection, with maxima of u and v being swept "downstream" as the component perpendicular to the ice edge increases. Surface stress, and particularly its curl, can be very sensitive to such changes.

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